AN EARLY-SEASON SNOWSTORM ALONG THE ATLANTIC COAST, DECEMBER 4–5, 1957

JOHN B. FUGE AND JOHN M. KIPPER, JR.

National Weather Analysis Center, U. S. Weather Bureau, Washington, D. C.

1. INTRODUCTION

The snowstorm along the Atlantic Seaboard on December 4–5, 1957, was unusual only in that it occurred so early in the season, being in other respects only a typical winter nor'easter. However, the snowfall, over one foot in places, made it one of December's outstanding meteorological features.

An attempt is made to use Petterssen's [1] development equation in a qualitative manner to account for the development of the storm together with a method of estimating divergence proposed by Cressman [2]. An estimate of the precipitation is made using a method presented by Swayne [3].

2. DEVELOPMENT

The development of the storm is investigated with the aid of Petterssen's equation for the local change in vorticity at the surface $\partial\eta_0/\partial t$. The equation may be written, neglecting terms shown to have negligible contribution (cf. Means [4]):

$$\frac{\partial\eta_0}{\partial t}=A_\eta-\frac{g}{f}\nabla^2A_h-\frac{R}{f}\nabla^2\left[\ln\frac{p_0}{p}\left(\overline{\omega(\Gamma_a-\Gamma)}+\frac{1}{c_p}\overline{\frac{dW}{dt}}\right)\right]$$

where:

A_η is the vorticity advection at the level of non-divergence.

$-\frac{g}{f}\nabla^2A_h$ is the development contribution of thickness advection for the layer 1000 mb. to the level of non-divergence; it is proportional to the Laplacian of thermal advection A_h for the layer. g is the acceleration of gravity and f is the Coriolis parameter.

$\overline{\omega(\Gamma_a-\Gamma)}$ is the stability, or "buoyancy" term, whose negative Laplacian represents the development contribution of local thickness changes due to adiabatic processes. $\omega\equiv dp/dt$, and Γ_a and Γ are respectively the adiabatic and actual lapse rates with respect to pressure.

$\frac{1}{c_p}\overline{\frac{dW}{dt}}$ is the mean rate of heating of the layer due to non-adiabatic processes. The negative Laplacian of this term therefore represents the development contribution of local thickness changes due to non-adiabatic processes.

c_p is the specific heat at constant pressure.

R is the gas constant, p_0 and p are the pressure respectively at 1000 mb. and at the level of non-divergence.

Since the terms were used only in a qualitative sense to determine their contribution to low-level convergence they were investigated separately even though they are not independent processes. For the same reason, after inspection of the relevant upper-level charts, it was felt that the assumption of an isobaric level of non-divergence at 500 mb. would not change the signs of the terms contributing appreciably to the vorticity tendency at the surface. It can be seen from the development equation that negative Laplacians of the thermal terms together with positive vorticity advection will give positive contributions to vorticity production and thus intensification at the surface.

To study the vorticity advection at the level of non-divergence ($A\eta$), the vorticity at 500 mb. was computed by using the finite difference method with a grid length of 200 km. The vorticity field was then superimposed on the 500-mb. contours and the vorticity advection was studied.

The Laplacian of thermal advection $-\frac{g}{f}\nabla^2A_h$ was studied by means of the 1000-500-mb. thickness charts. The field of advection was computed as the product of the geostrophic wind component and the thickness gradient measured over a distance of 100 nautical miles. Multiplication by a suitable factor gave the result as thermal advection in °C. per 12 hours. The field of thermal advection was then plotted and the Laplacian computed by finite differences over a 200-km. grid. To define the areas of positive or negative Laplacian, 25 to 40 points were used at each observation time.

The buoyancy term $-\frac{R}{f}\nabla^2\left[\ln\frac{p_0}{p}\overline{\omega(\Gamma_a-\Gamma)}\right]$ and the term for the contribution of local thickness changes due to non-adiabatic processes, $-\frac{R}{f}\nabla^2\left[\ln\frac{p_0}{p}\frac{1}{c_p}\overline{\frac{dW}{dt}}\right]$ were only qualitatively evaluated.

As vertical motion charts were not being routinely prepared by the Joint Numerical Weather Prediction (JNWP) Unit at this time, the sense of vertical motion was inferred from the divergence, which was qualitatively estimated by using the method proposed by Cressman [2]. Essentially this utilizes the vorticity equation with certain assumptions[1] to obtain the equation

$$\nabla_H \cdot \mathbf{V} = \mathbf{V}_T - \nabla_H \ln \eta$$

where $\nabla_H \cdot \mathbf{V}$ is the horizontal divergence of the wind vector $\mathbf{V}$ at the level in question, $\mathbf{V}_T$ is the vector wind difference between the level of non-divergence and the level in question, and η is the absolute vorticity (vertical component) at the level where the divergence is desired. This logarithmic vorticity advection, which is inversely proportional to the size of the areas formed by intersections of thickness lines with vorticity lines at the level in question, was not actually computed. It was visually estimated from superimposing the thickness and vorticity charts. Thus, if the thermal wind blows from low to high vorticity at the lower level, divergence exists; or if the thermal wind blows from high to low vorticity, convergence exists.

This was correlated with the stability index charts and it was found that in general the maximum contribution of the stability term was in the areas of maximum warm air advection. This is, of course, to be expected since only in areas where the vorticity lines were not parallel to the contours would significant differences have occurred. While the pattern of the stability index (fig. 10) showed some change through the period, with less positive values following the developing surface system, the values did remain large and positive. Thus the buoyancy term would act as a brake over the land areas. Offshore it is likely that, with the aid of the non-adiabatic heating term, the stability could be negative and contribute to the development.

Indications that the entire development took place on the Arctic front were evident by near or below normal values of the departure from normal 1000–500-mb. thicknesses south of the front and eastward into the Atlantic during the period considered. This was the case not only in the upper part of the 1000–500-mb. layer, but was also apparent from inspection of the surface elements and the fact that the amounts of precipitation occurring in the Southeastern States during this period were not those associated with tropical air.

As for the synoptic events, earlier charts (not shown) showed that on December 2 a Pacific occlusion associated with a 500-mb. short-wave trough moved eastward into the Texas Panhandle where a wave became organized on the Arctic front that lay in an essentially east-west line along the 35th parallel. By 0000 GMT, December 4 (fig. 1) the wave had moved into Kentucky and deepened 5 mb. However, as can be seen from figure 4, the axis of maximum vorticity advection was already east and south of the surface Low, indicating the development over the center should be at a decreasing rate, especially if the favorable Laplacian of thermal advection over West Virginia (fig. 7) should be diminished. On the other

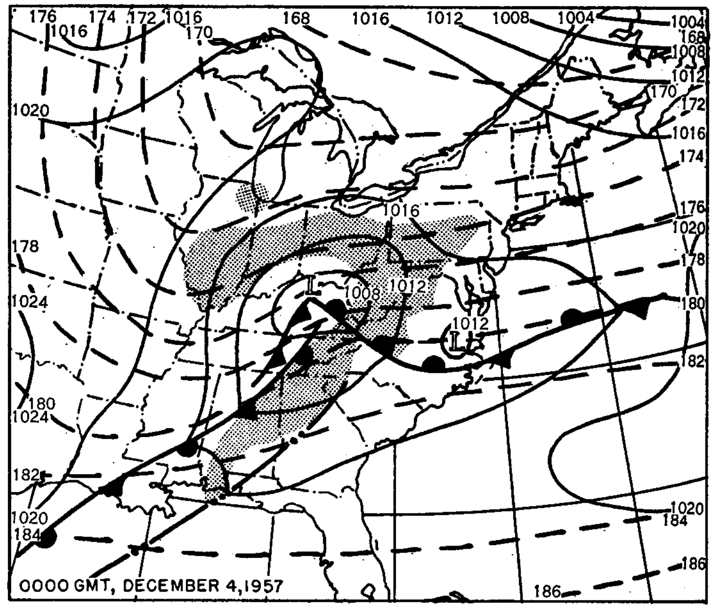

FIGURE 1.—Sea level chart for 0000 GMT, December 4, 1957, with 1000–500-mb. thickness (dashed lines). Shaded areas indicate current precipitation.

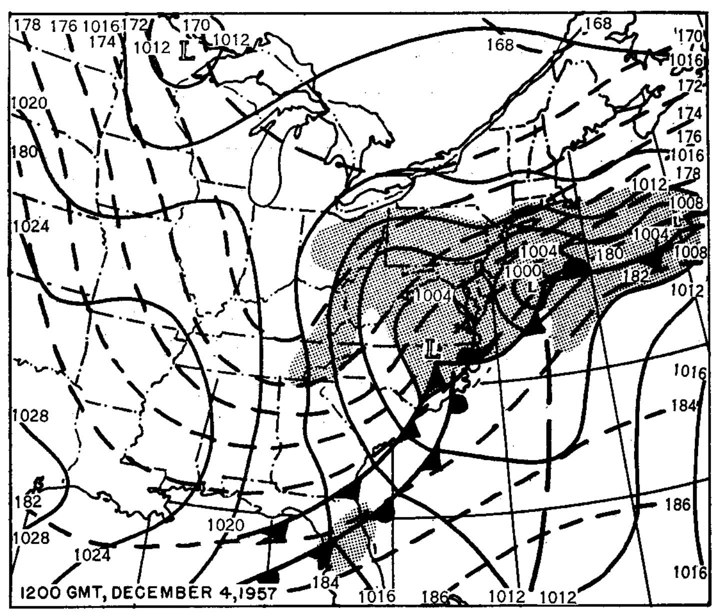

FIGURE 2.—Sea level chart for 1200 GMT, December 4, 1957, with 1000–500-mb. thickness (dashed lines). Shaded areas indicate current precipitation.

hand, along this axis of maximum vorticity advection from Pennsylvania to Alabama (fig. 4) there was a small area of much larger values in eastern Tennessee superimposed upon a near zero value of the Laplacian of thermal advection. According to Petterssen this should be a favored area for development, especially if there were any contributions from the other terms. Positive vorticity advection aloft, just east of Norfolk, was also favorable for intensification in that area. The lesser magnitude of the vorticity advection was probably compensated for by a strong contribution to vorticity from the non-adiabatic term as a result of heating from the underlying

[1] Neglecting horizontal solenoids, friction, vertical transport of vorticity, and rotation of vorticity from about a horizontal to about a vertical axis.

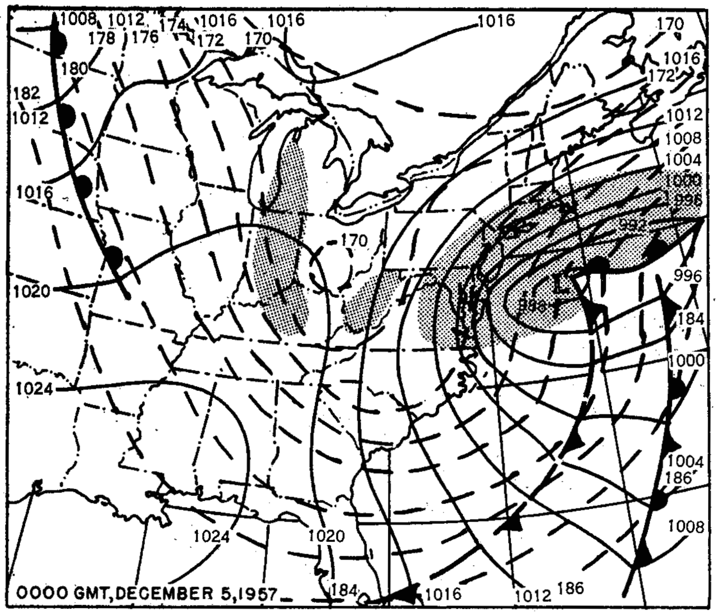

FIGURE 3.—Sea level chart for 0000 GMT, December 5, 1957, with 1000–500-mb. thickness (dashed lines). Shaded areas indicate current precipitation.

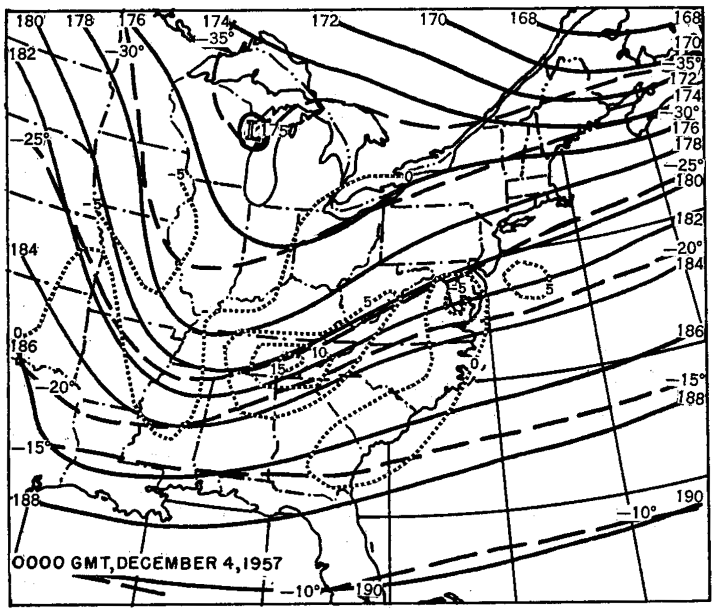

FIGURE 4.—500-mb. contours and isotherms (dashed lines) for 0000 GMT, December 4, 1957. Vorticity advection indicated by dotted lines in units 10^{-9} sec^{-2}.

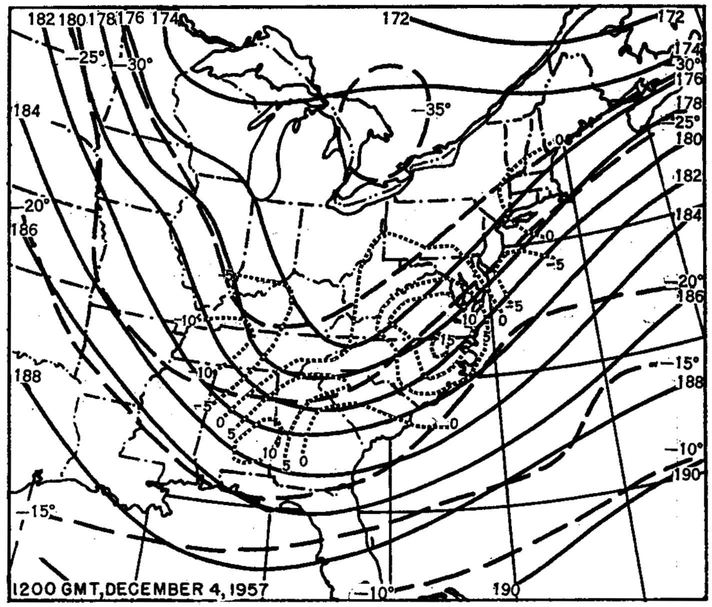

FIGURE 5.—500-mb. contours and isotherms (dashed lines) for 1200 GMT, December 4, 1957. Vorticity advection indicated by dotted lines in units 10^{-9} sec^{-2}.

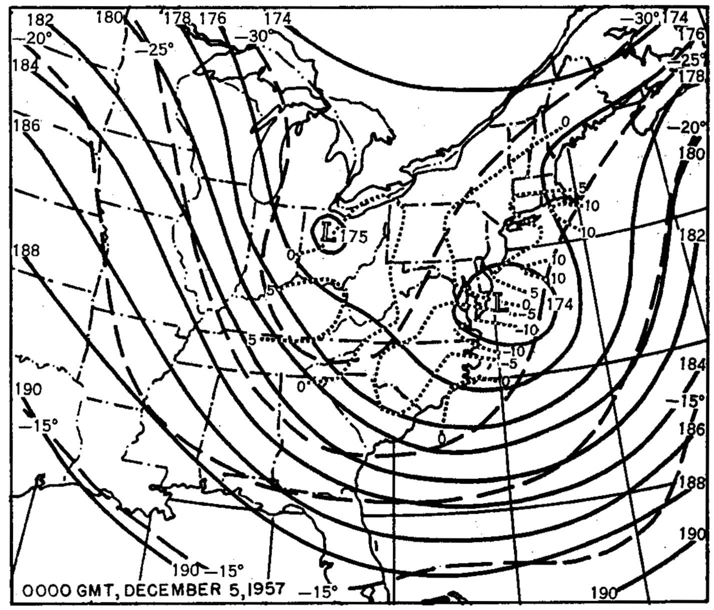

FIGURE 6.—500-mb. contours and isotherms (dashed lines) for 0000 GMT, December 5, 1957. Vorticity advection indicated by dotted lines in units 10^{-9} sec^{-2}.

sea surface. This heating effect would in turn react favorably on the stability term. Thus maximum intensification should occur on an axis from east of Norfolk through eastern Tennessee.

By 1200 GMT, December 4 (fig. 2) the maximum intensification did occur along that axis and two Lows, one in North Carolina and the other out to sea, developed and were both increasing in intensity. With no vorticity advection and rapidly decreasing temperature contributions, the original Low over Kentucky moved into eastern West Virginia and filled. Even with the area of maximum vorticity advection aloft directly overhead (fig. 5), indicating its maximum contribution had already been made, the Low in North Carolina continued to intensify as it moved offshore and received the contribution of non-adiabatic heating from the sea surface. The final map, 0000 GMT, December 5 (fig. 3), shows that the two Lows merged into a large east-west trough which was still increasing in intensity. With the Low aloft now much closer to the surface Low, the vorticity advection

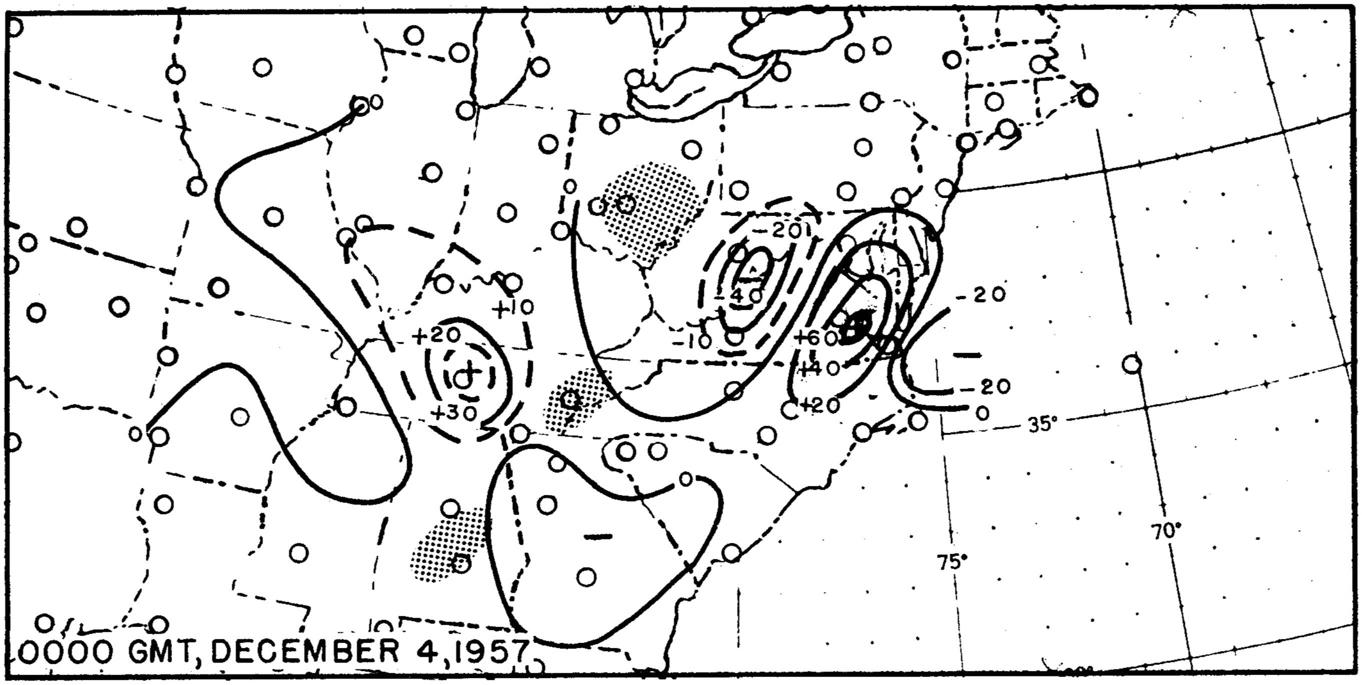

FIGURE 7.—Laplacian of thermal advection computed from surface geostrophic winds and 1000–500-mb. thickness, in units 10^{-9} sec^{-2} Shaded areas represent 6-hour precipitation amounts greater than 0.25 inch ending at 0000 GMT, December 4, 1957.

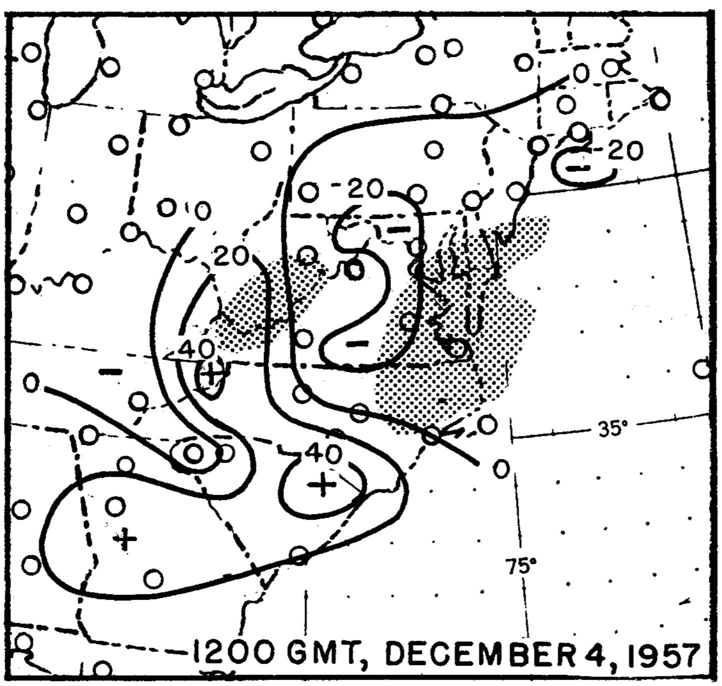

FIGURE 8.—Laplacian of thermal advection computed from surface geostrophic winds and 1000–500-mb. thickness, in units 10^{-9} sec^{-2}. Shaded areas represent 6-hour precipitation amounts greater than 0.25 inch ending at 1200 GMT, December 4, 1957.

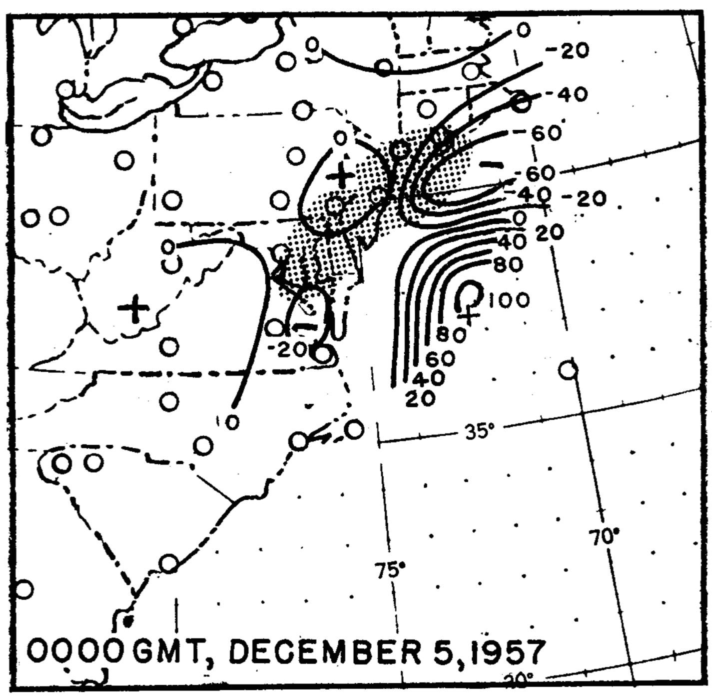

FIGURE 9.—Laplacian of thermal advection computed from surface geostrophic winds and 1000–500-mb. thickness, in units 10^{-9} sec^{-2}. Shaded areas represent 6-hour precipitation amounts greater than 0.25 inch ending at 0000 GMT, December 5, 1957.

(fig. 6) had already decreased appreciably over the western portion of the center, but a large negative Laplacian of thermal advection along the coast (fig. 9) was still contributing appreciably to intensification. This intensification did continue after 0000 GMT, but at a decreasing rate.

It is of interest to note that comparison of the Laplacian of thermal advection with the areas of precipitation (figs. 7, 8, 9) shows approximate agreement when the motion of the systems is considered. This is in line with the finding of other investigators (cf. [4]). Some irregularities in the pattern of the Laplacian of thermal advection over the

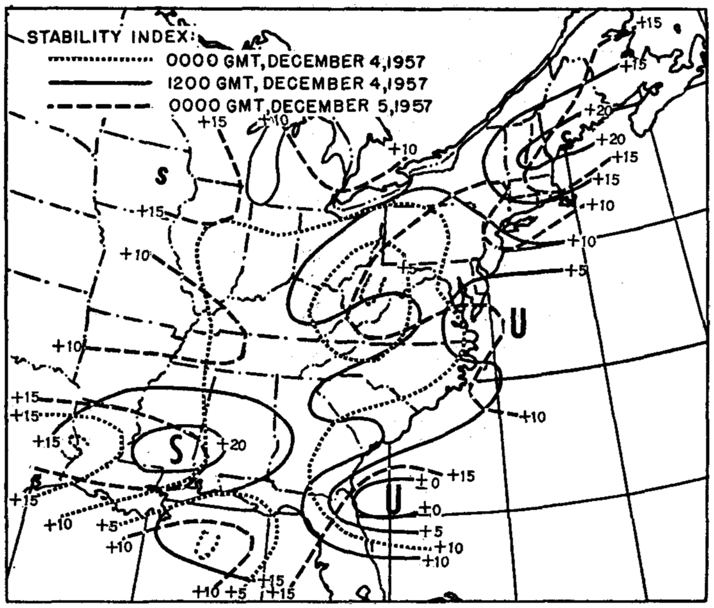

FIGURE 10.—Stability index chart.

mountains were due to the topography and thus would not be associated with precipitation since moisture was not available.

3. PRECIPITATION COMPUTATIONS

In an article of this scope only an estimate of precipitation could be made. Lacking even quantitative values of vertical motion, the best approach was one using only the data at hand. In addition, lack of data over the ocean area precluded anything but a simple approach. Accordingly, a method proposed by Swayne [3] was utilized. He has shown that with certain simplifying assumptions the precipitation rate is proportional to the moisture advection. First it is necessary to convert the precipitable water to "saturation thickness." This is defined as the thickness between specified constant pressure surfaces of a saturated column having the same precipitable water value as the observed column. It is then assumed that the thickness lines in saturated areas are fixed for the duration of the precipitation. In this case, as has been found elsewhere, the changes were negligible—at least in the area where computations were made for the periods concerned. Then, for the special case of unsaturated air, using the mid-layer geostrophic wind and ignoring vertical moisture transport other than that implied by the fact that the thickness lines are not advected in the precipitation areas, it can be shown, with some further simplifications, that the rate of precipitation is proportional to the area formed by the grid of thickness lines and contours.

Computations were made for each of the four 6-hour periods beginning at 0000 GMT on December 4 and continuing through 0000 GMT, December 5, using the 1000–700-mb. thickness and the 850-mb. contours. Since the contour field could not be drawn for 0600 and 1800 GMT, the actual winds were used. The difference between actual contours and contours implied by the winds is, in this case, within the limitations of the method itself. To obtain a more representative verification, the precipitation amounts were taken 3 hours either side of the computation times. Results are shown in table 1, for the Washington, Baltimore, and New York stations where hourly values and intermediate winds aloft were available. Some of the larger discrepancies could be qualitatively accounted for.

The computation for 0000 GMT, December 4 at Washington was closely approximated 40 miles southeast at

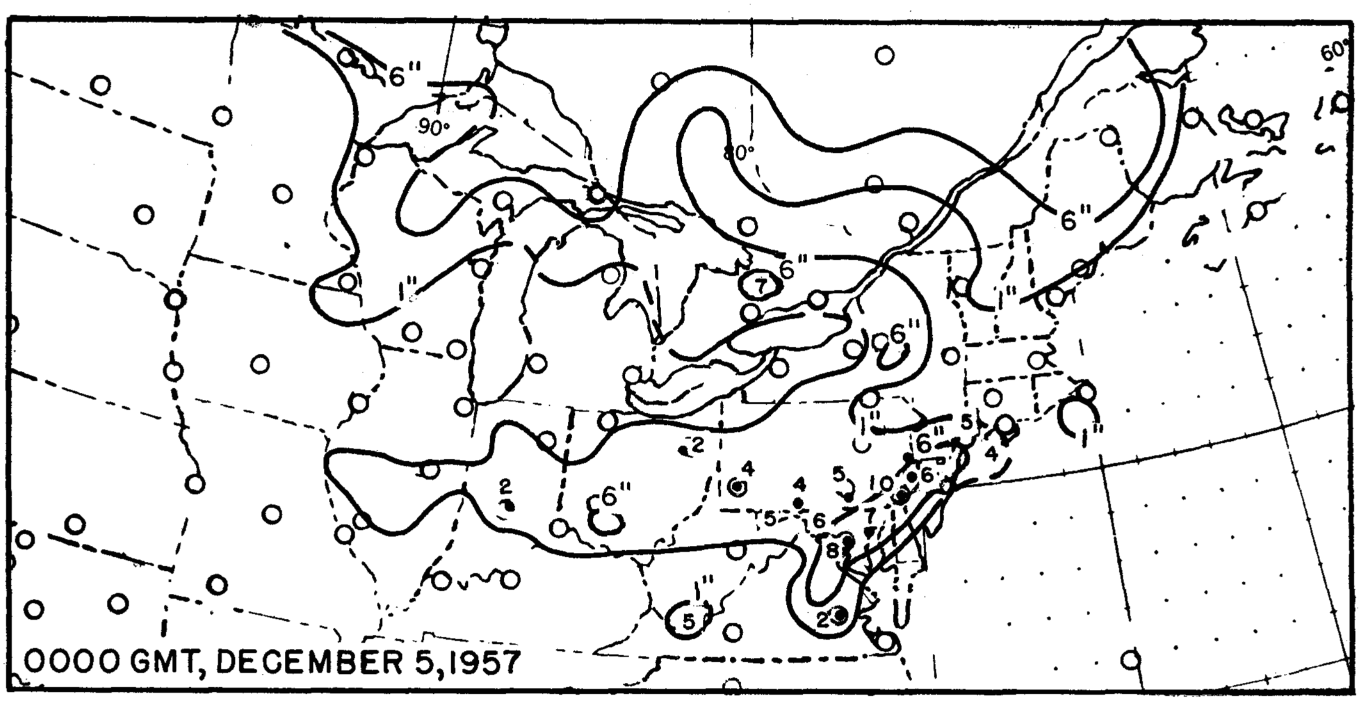

FIGURE 11.—Snow depth on the ground (inches) at 0000 GMT, December 5, 1957.

TABLE 1.—*Computed and observed 6-hour precipitation amounts*

Date GMT	Computed	Observed				Computed	Observed		
		Washington, D. C.		Baltimore, Md.			New York, N. Y.		
		City office	Airport	City office	Airport		City office	La Guardia	Central Park
Dec.									
4th 0000	0.24	0.02	0.02	0.03	0.03	0.03	0.01	0.02	T
4th 0600	.18	.17	.04	.17	.10	.35	.24	.18	.20
4th 1200	.00	.31	.26	.29	.21	.36	.13	.09	.05
4th 1800	.27	.42	.62	.31	.25	.48	.36	.40	.39
5th 0000	.00	.08	.19	.03	.07	.15	.15	.26	.10
Total	.57	1.00	1.13	.83	.66	1.37	.89	.95	.74

Patuxent, Md., which reported .33 inch in the period ending at 0600 GMT, or 6 hours after computation time. This value seemed fairly representative of the area just to the south. At 0000 GMT on the 5th, in connection with the computed and observed values, the precipitation stopped within one hour after the computation time in the Washington-Baltimore area. At 1200 GMT on the 4th, the 850-mb. Low was centered directly over the Washington-Baltimore area but the Low centers at higher levels were still to the west. Inspection of the Washington winds showed no appreciable cross-thickness flow at any level below the 700-mb. level. Computations for a deeper layer were not made as there were considerable variations in the hourly amounts. While city office values for both Washington and Baltimore were evenly distributed throughout the period, bracketing the computations, both airports showed that nearly all of the precipitation, .20 inch in each case, fell in the last 2 hours. Surface winds were still light at this time so that varying exposures do not account for the difference. Also at 1200 GMT, in the New York area, inspection of the upper winds showed apparent cold advection between 12,000 and 20,000 feet. It is possible that the implied variation in vertical velocity affected the rate of precipitation.

It is difficult to assess the effect of varying exposures as there were differences between airports and downtown offices during periods of light winds and similarities during periods of strong winds. Also, as Swayne [3] has shown, factors which could introduce sizable errors are: ageostrophic motions, non-advective processes, and changing wind patterns. In addition, errors were introduced by the neglect of those scales of the processes which could not be investigated because of limitations of the data. Considering these factors and the assumptions made, the overall agreement between computed and observed values was very good.

4. PROGNOSTIC PROBLEMS

The JNWP 500-mb. barotropic prognostic charts for 12, 24, and 36 hours verifying at 0000 GMT, December 5, while progressively improving, still underestimated the trough development along the Atlantic coast. It would appear that the baroclinicity must have been the dominant cause of error here. Kinematic techniques from the 1200 GMT charts of December 3 were also in error which might be expected in a developing system. The best prognostic results at that time were obtained using a combination of extrapolation and the control line technique [5].

5. CONCLUSION

In conclusion then, the original Low center dissipated in West Virginia with formation of a new center offshore. Contrary to the usual sequence, in which the offshore Low immediately becomes dominant, formation of a new, intense center took place in a climatologically infrequent location—extreme northwestern North Carolina—as the area of maximum vorticity advection became superimposed on a favorable thermal field. This new center moved offshore and became the dominant Low, only slowly beginning to merge with the first offshore center after precipitation stopped in the area of heaviest snowfall. Rates of snowfall were light but the intensification of the second system allowed snow to continue 24 hours in places, with resulting large accumulations.

ACKNOWLEDGMENTS

The authors wish to thank the staff of the Daily Map Unit for drafting the illustrations and members of NAWAC for helpful suggestions and assistance.

REFERENCES

1. S. Petterssen, *Weather Analysis and Forecasting*, 2d ed., McGraw-Hill Book Co., Inc., New York, 1956.
2. G. P. Cressman, "An Approximate Method of Divergence Measurement," *Journal of Meteorology*, vol. 11, No. 2, Apr. 1954, pp. 83–90.
3. W. W. Swayne, "Quantitative Analysis and Forecasting of Winter Rainfall Patterns," *Monthly Weather Review*, vol. 84, No. 2, Feb. 1956, pp. 53–65.
4. L. L. Means, "An Investigation of Cyclone Development—Storm of December 13–15, 1951," *Monthly Weather Review*, vol. 83, No. 9, Sept. 1955, pp. 185–198 (eq. (2.6), p. 197).
5. P. E. Wasko, "The Control-Line Method of Constructing Prognostic Charts," *Bulletin of the American Meteorological Society*, vol. 33, No. 6, June 1952, pp. 233–236.